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Archean and Paleoproterozoic fault history of the Big lake shear zone, MacQuoid–Gibson lakes area, Nunavut[1]

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Abstract: The Big lake shear zone in the MacQuoid–Gibson lakes area comprises diverse mylonitic rocks with a complex tectonothermal history. The shear zone has an exposed strike length of 50 km, with abrupt termination to the west. Central and eastern segments of the shear zone consist of dextral, amphibolite-facies, porphyroclastic mylonite and annealed straight gneiss. The western segment, however, is cored by variably retrogressed granulite-facies mafic and anorthositic ultramylonite, within amphibolite-facies wall rocks, and has a more complex kinematic history. The granulite-facies rocks are spatially associated with the anorthosite sheets, which are interpreted here as the dominant heat source of the localized high-grade event. The ultramylonite units are crosscut by undeformed ca. 2.19 Ga MacQuoid dykes, constraining the high-grade, high-strain event to be older. Regional aeromagnetic data illustrate that the western segment is truncated by two faults of (?)Proterozoic age.

Résumé : La zone de cisaillement de Big Lake dans la région des lacs MacQuoid et Gibson comporte une gamme de roches mylonitiques ayant une histoire tectonothermique complexe. La zone de cisaillement est exposée parallèlement à sa direction sur 50 km et se termine brusquement à l'ouest. Ses segments central et oriental sont constitués de mylonite porphyroclastique du faciès des amphibolites ayant subi un cisaillement dextre et de gneiss rectiligne recuit. Par contre, le segment occidental contient un noyau d'ultramylonite anorthosique et mafique variablement rétromorphosé du faciès des granulites au sein d'épontes du faciès des amphibolites et son une histoire cinématique est plus complexe. Les roches du faciès des granulites sont associées dans l'espace aux nappes d'anorthosite, lesquelles sont interprétées dans le présent article comme étant la principale source de chaleur pour le métamorphisme intense local. Les unités d'ultramylonite sont recoupées par les dykes de MacQuoid non déformés (environ 2,19 Ga), ce qui signifie que l'événement ayant produit le métamorphisme intense et la forte contrainte est plus ancien. Des données aéromagnétiques régionales indiquent que deux failles (?)protérozoïques coupent le segment occidental.

[1] Contribution to the Western Churchill NATMAP Project

INTRODUCTION

The Gibson Lake–Cross Bay–MacQuoid Lake area (Fig. 1) comprises a multiply deformed composite terrane of Archean metavolcanic, metasedimentary, and diverse plutonic rocks, all tectonothermally reworked and intruded by plutons and dyke swarms in the Paleoproterozoic (*see* Hanmer et al., 1999a, b; Tella et al., 1999, 2000). The area (Fig. 1) is divided into 1) a southeastern panel of northwest-dipping metasedimentary rocks and gneissic tonalite sheets (MacQuoid–Gibson homocline); 2) an overlying metavolcanic belt to the north and west, which hosts several tonalite-granodiorite plutons; 3) the Cross Bay plutonic complex in the northeast, comprising gneissic tonalite and diorite-gabbro intruded by younger granite bodies; and 4) the Big lake shear zone, which transects the Archean gneiss units, and comprises a zone of annealed straight gneiss and ribbon mylonite units. This report focuses on the anatomy and deformation history of the Big lake shear zone.

Paleoproterozoic intrusive rocks that bracket the timing of deformation along the Big lake shear zone include the ca. 2.19 Ga MacQuoid dyke swarm (Tella et al., 1997), ca. 1.83 Ga lamprophyre dykes (MacRae et al., 1996), and ca. 1.83 Ga isotropic granite bodies (Tella et al. 1997; W. Davis, unpub. data, 1999).

The Big lake shear zone was previously thought to be mostly a Paleoproterozoic structure (Hanmer et al., 1999a; Ryan et al., 1999) because ca. 2.19 Ga MacQuoid dykes within and adjacent to the central segment of the Big lake shear zone are boudined; but the age of initiation of the shear zone was unconstrained. During the 1999 fieldwork, however, undeformed MacQuoid dykes were observed cutting across ultramylonitic rocks in the western segment of the Big lake shear zone, convincingly demonstrating that the ultramylonite formed prior to ca. 2.19 Ga. We interpret the boudinage of MacQuoid dykes in the east to reflect subsequent regional deformation not recorded in the western segment. On the other hand, fault structures which appear to truncate the Big lake shear zone beneath the Quaternary cover appear to be associated with formation and modification of the ca. 1.83 Ga Baker Lake Basin.

BIG LAKE SHEAR ZONE

The Big lake shear zone is a generally east-striking, steeply dipping zone of intensely sheared rocks that broadly coincides with the southern boundary of the Cross Bay complex (Fig. 1; Hanmer et al., 1999a, b). Based on lithological and structural differences along strike, the shear zone is divided into three segments (eastern, central, and western).

Central and eastern segments

The central and eastern segments of the shear zone are described in Ryan et al. (1999) and Hanmer et al. (1999b), and the main characteristics are summarized here. These segments mainly comprise amphibolite-facies mylonite derived from a tonalitic protolith, metasedimentary inclusions, and abundant sheets of intensely sheared monzogranite. The shear zone has zones of porphyroclastic, relatively coarse-grained annealed straight gneiss, and zones of finer grained ribbon mylonite. The annealed straight gneiss is locally difficult to distinguish from the regional foliation in orthogneissic and paragneissic wall rocks, except that the straight gneiss units are consistently steeply dipping, are porphyroclastic, and have a well developed, subhorizontal, east- or west-plunging extension lineation.

Winged porphyroclasts in the central and eastern segments are composed primarily of quartz and feldspar (porphyroclasts of garnet, hornblende, and clinopyroxene are also locally abundant), and are predominantly symmetrical. The asymmetrical variety is dominated by delta geometry (*see* Fig. 8, Ryan et al., 1999; Fig. 7, 8, Hanmer et al., 1999b), indicating dextral shear along the Big lake shear zone. Other dextral shear-sense indicators include z-asymmetric shear drag folds, asymmetrically pulled-apart boudins of amphibolite layers (*see* Fig. 9, Ryan et al., 1999), oblique shape fabrics, and shear bands. Hanmer et al. (1999b) and Ryan et al. (1999) interpreted the fabric elements and shear-sense indicators in the central and eastern segments as products of a dextral strike-slip shear zone of monoclinic symmetry.

Western segment

The western segment of the Big lake shear zone is distinct from the central and eastern segments because 1) it preserves granulite-facies assemblages, 2) it encompasses a wider compositional variety of rocks, 3) the orientations of its fabric elements are different than those in the east, 4) shear-sense indicators show a more complex deformation history, and 5) the crosscutting nature of various intrusive bodies better constrains the timing of early displacement along this part of the shear zone. Rocks in the western segment, in order of decreasing abundance, include tonalitic orthogneiss, homogeneous tonalite, paragneiss, augen monzogranite (ca. 2.69 Ga; W. Davis, unpub. data, 1999), mafic granulite (variably retrogressed to garnet-amphibolite), anorthosite, and diabase.

Proterozoic monzogranite and coeval lamprophyre units (Sandeman et al., 2000) crosscut the mylonitic rocks. An approximately 20 km^2 monzogranite pluton, the "Squiggly lake" pluton (Fig. 2), crosscuts and effectively stitches the mylonite units of the Big lake shear zone (Hanmer et al., 1999a). It is a composite monzogranite pluton, with subordinate phases varying from hornblende-phlogopite gabbro to monzonite (Sandeman et al., 2000). It yielded a preliminary U/Pb of zircon age of ca. 1.83 Ga (W. Davis, unpub. data, 1999).

Despite preserving the highest grade rocks in the area, the western segment contains mylonite units that have the finest grain size of any in the map area. This character appears to be independent of rock type, because examples of aphanitic mylonite derived from monzogranite, anorthosite, diabase, and mafic rocks of uncertain protolith occur together. Here we focus on the mafic and anorthositic rocks.

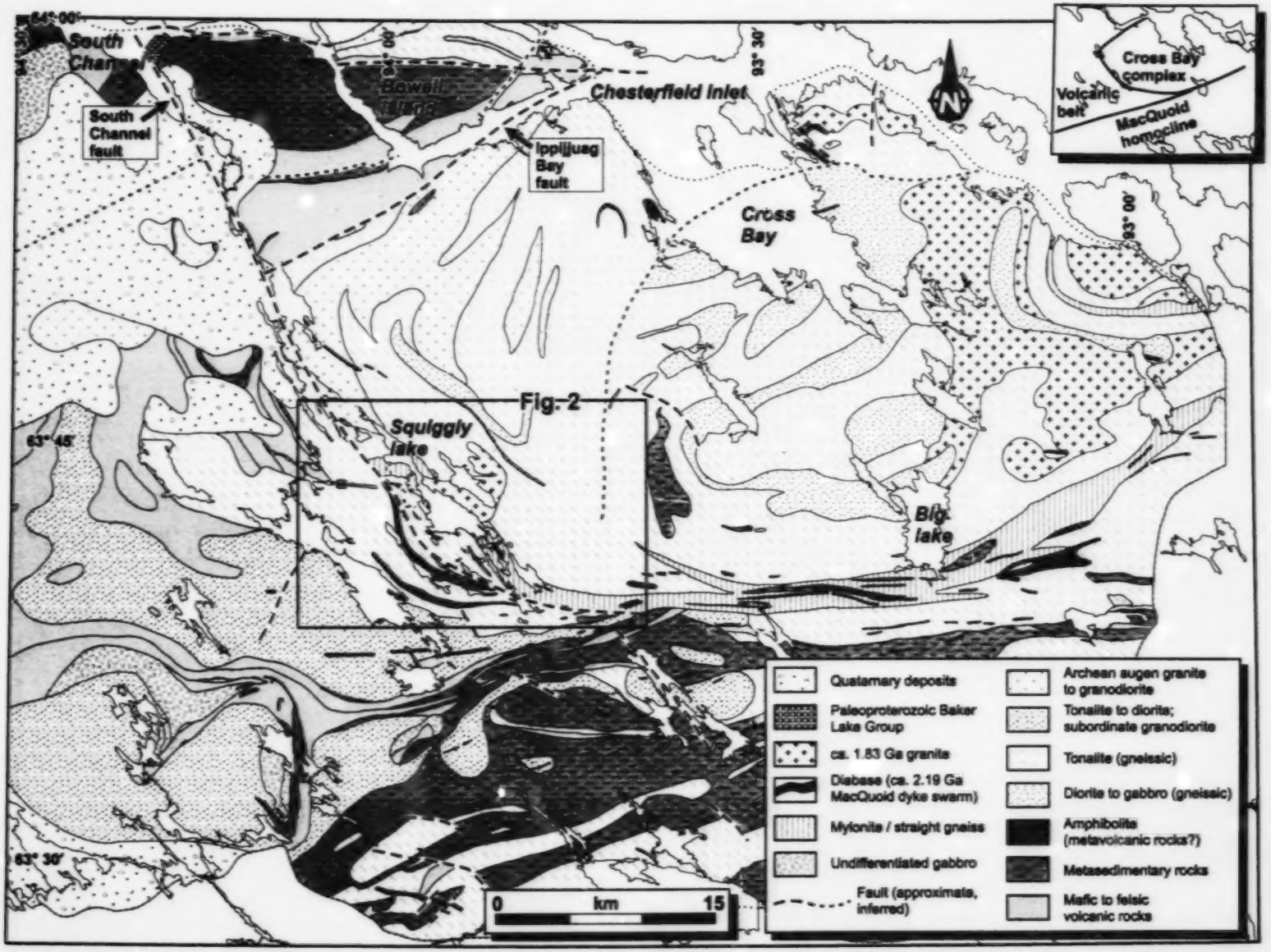


Figure 1. *Generalized map of the Gibson Lake–Cross Bay–MacQuoid Lake area* (after *Tella et al., 1999*). *The location of Figure 2 is outlined. Note the location of the South Channel fault and the Ippijjuaq Bay fault.*

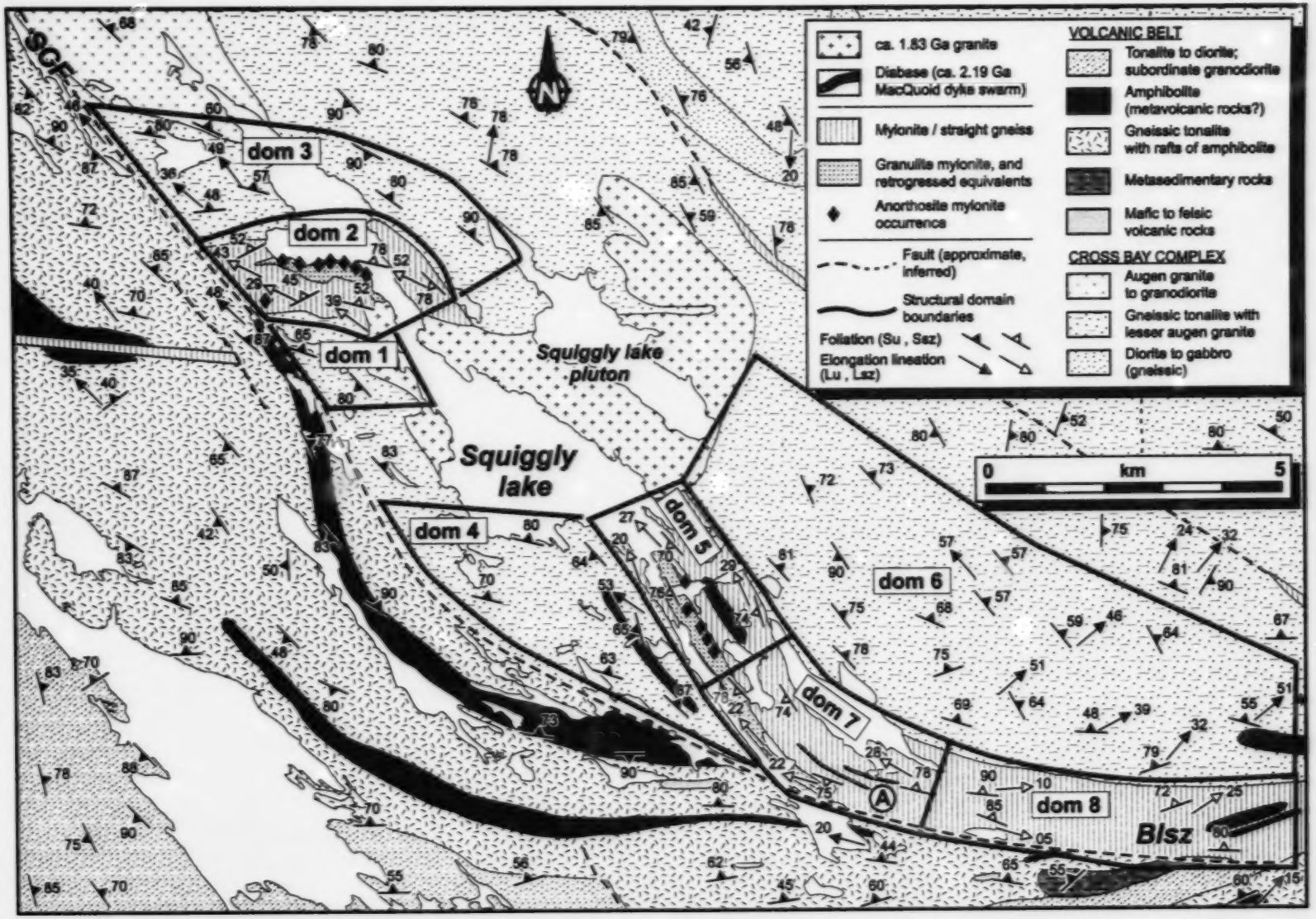


Figure 2. *Detailed structural map of the Squiggly lake area, outlining the Big lake shear zone (Blsz), and the Squiggly lake pluton. The Big lake shear zone is truncated by the South Channel fault (SCF). Eight structural domains (dom=domain) are outlined based on orientation and strain state of fabric elements. Note the distribution of granulite-facies and anorthositic mylonite units. Su=regional foliation; Ssz=shear foliation; Lu=regional elongation lineation; Lsz=shear zone extension lineation.*

Anorthositic mylonite units were noted during the 1998 mapping (Ryan et al., 1999; Quigley, 1999), and were thought to represent less than 10%, by volume, of the rocks in the western segment. This study confirms this low estimate, and shows that the most extensive and best preserved anorthosite sheets lie within the shear zone immediately north of the Squiggly lake pluton (Fig. 2), where some sheets are as thick as 10 m (Fig. 3a). Less abundant, thinner anorthosite sheets were mapped at the south end of Squiggly lake (Fig. 2). Because they are mylonitized and contain multiple generations of intrafolial isoclinal folds, the original thickness is unknown. They contain as much as 95% plagioclase and thin (1–8 mm) mafic layers (*see* Fig. 11, Ryan et al., 1999). Rare low-strain domains demonstrate that their protolith was gabbroic anorthosite. The mafic layers appear to have been derived from the mechanical breakdown and transposition of primary pyroxene, and are now composed of small (<1 mm), randomly oriented, acicular, actinolitic hornblende. The random orientation of these amphibole minerals indicates that they grew after deformation, and that hydration must have occurred locally in these rocks. The preservation of granulite demonstrates that the hydration was not pervasive. In contrast, plagioclase layers have remained remarkably fine grained (<50–100 μm).

High-grade mylonite units were locally recognized during the 1998 mapping (Ryan et al., 1999). Follow-up petrographic investigations (e.g. Quigley, 1999) and this year's mapping demonstrate that granulite-facies mylonite units are much more extensive than originally thought, extending for 10 km, along the entire northwest-striking portion of the western segment (Fig. 2). The best preserved high-grade assemblages are in mafic rocks, and comprise garnet-orthopyroxene-clinopyroxene-plagioclase±hornblende±quartz±titanite±zircon. This assemblage is present in both the coarse-grained gneissic granulite rocks (Fig. 3b), and the ribbon ultramylonite rocks (Fig. 3c). Preliminary pressure-temperature estimates indicate high-pressure, granulite-facies metamorphic conditions of about 800°C and 12 kbars (R. Berman, unpub. data, 1999). Some specimens contain abundant porphyroblasts of zircon, titanite, rutile, and monazite. The mafic granulite units are commonly postkinematically retrogressed to coarse-grained

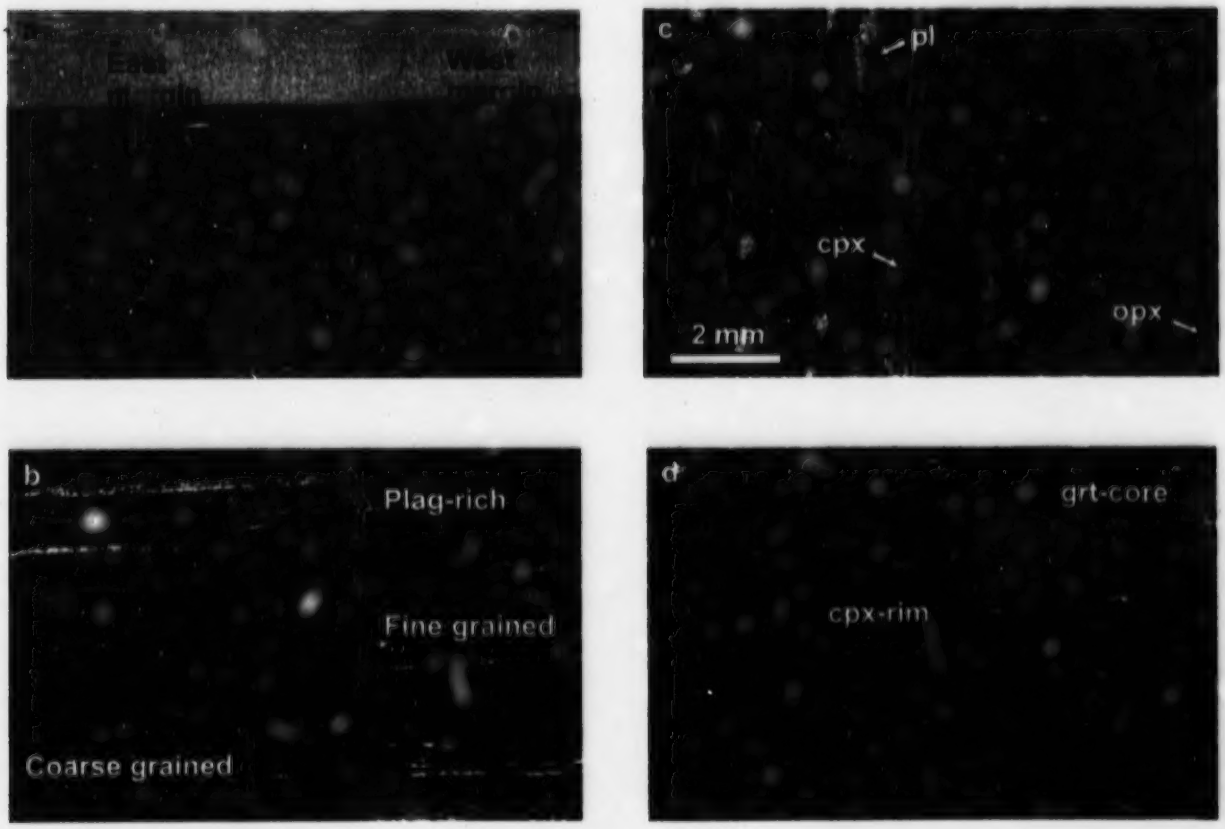
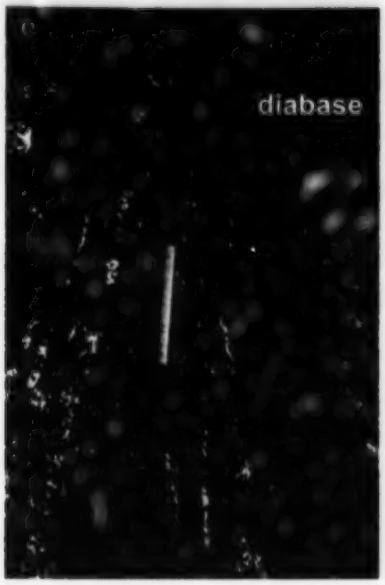
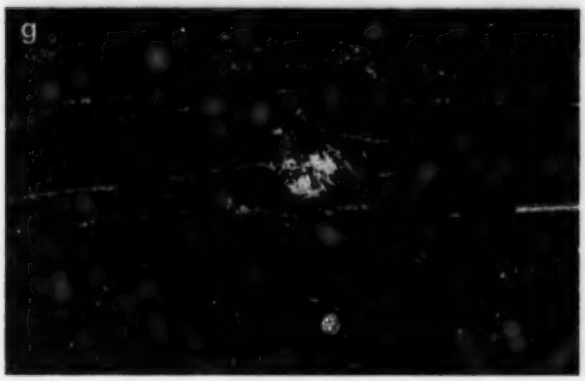


Figure 3. a) Ten metre thick layer of anorthositic mylonite (light coloured) in the Big lake shear zone. b) Compositionally layered mafic granulite gneiss, of variable grain size, in the Big lake shear zone; Plag=plagioclase. c) Photomicrograph of mafic granulite ribbon ultramylonite. Garnet porphyroclasts are smaller that clinopyroxene (cpx), orthopyroxene (opx), and plagioclase (pl). d) Composite garbet-clinopyroxene porphyroclast in coarse-grained granulite mylonite.





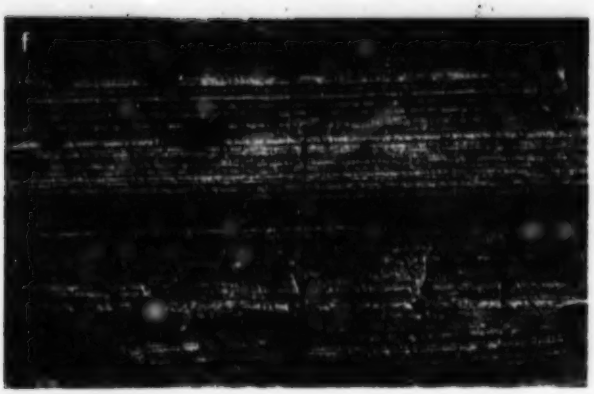


Figure 3. e) Late-synkinematic diabase obliquely cutting mylonite, and is itself mylonitized. f) Mafic granulite ultramylonite. g) Sinistral delta-winged felspar porphyroclast in amphibolite mylonite in Big lake shear zone west of Squiggly lake. h) Dextral shear-drag fold of quartz vein in the Ippijjuaq Bay fault zone.

garnet-hornblende-biotite-plagioclase (±orthopyroxene± clinopyroxene±quartz) amphibolite gneiss and schist. Composite porphyroclasts of garnet-clinopyroxene are commonly more than 5 cm (Fig. 3d). Due to the degree of metamorphism and/or strain, the protolith of the mafic granulite remains equivocal. It is possible that some mafic layers represent gabbroic sheets, perhaps magmatically associated with the anorthosite. However, sheets with gabbroic textures have not been mapped in the wall rocks or low-strain domains. The best-preserved granulite-facies assemblages occur adjacent to the anorthosite sheets, suggestive of a thermal association between them. Granulite-facies assemblages die out progressively away from the anorthosite bodies, until only amphibolite-facies assemblages are present in the mafic lithologies in the wall rocks of the shear zone. These amphibolite units are demonstrably derived from rafts of volcanic country rock to the tonalite-monzogranite intrusions. We thus interpret the mafic granulite to be largely derived from mafic wall rocks (metavolcanic rocks), that have been thermally perturbed due to residual heat from the anorthositic and possible gabbroic intrusions.

A suite of fine-grained diabase dykes appears to have been intruded late-synkinematically into the shear zone (Ryan et al., 1999). These dykes are relatively abundant, straight walled, generally 0.20–2 m thick, and broadly parallel to the shear zone walls. In general, they are penetratively mylonitized, but locally preserve a 5–10° counter-clockwise obliquity to the mylonite foliation (Fig. 3e). They are strongly metamorphosed, comprising as much as 30 modal per cent pin-head size garnet. Based on these field characteristics, Ryan et al. (1999) speculated that these dykes are distinct from the MacQuoid dykes, which has been substantiated by geochemical and isotopic studies (H. Sandeman, unpub. data, 1999).

Although synkinematic intrusion of magma is, in general, difficult to prove, most observations associated with these diabase bodies are consistent with synkinematic localization of mafic magma in the shear zone. The diabase bodies are only recognized within or immediately adjacent to the shear zone. Intrafolial folds of wisps of mylonitized tonalite inclusions within the internally mylonitized diabase are parallel to the extension lineation in the shear zone. The metamorphic

mineral assemblage in the dykes is garnet-hornblende-plagioclase-biotite (with relict clinopyroxene). Larger (200–500 μm) relict garnets have low-calcium cores, and are wrapped by biotite in the mylonitic foliation. The biotite foliation is axial planar to intrafolial folds, and is overgrown by tiny (about 10–50 μm) calcium-rich garnets. Preliminary P-T estimates (R. Berman, unpub. data, 1999), using the younger garnets, indicate peak metamorphic conditions of upper amphibolite- to lower granulite-facies, high-pressure metamorphism (about 700–750°C, and 10–12 kbars). These temperatures were 50–100°C lower than the peak temperature of the host granulite mylonite (*see* above). Dykes in the adjacent wall rocks are also penetratively mylonitized, but the wall rocks themselves only show the ambient regional deformation; an apparent contradiction of the relative competence of diabase and tonalite. We consider that these characteristics are best explained by mafic magma having been intruded after mylonitization had begun, such that the diabase deformed in the ongoing event at elevated subsolidus temperatures while still under crystal plastic conditions.

Quartzofeldspathic mylonite units show variable states of postkinematic coarsening of grain size. They may contain abundant large garnet porphyroblasts, an uncommon observation regionally. They appear to have been susceptible to fluid infiltration after deformation, because much of the garnet was pseudomorphed by chlorite after deformation ceased.

The ca. 2.19 Ga MacQuoid dykes (Tella et al., 1997) crosscut the Neoarchean foliation throughout the map area and are themselves affected by two episodes of Paleoproterozoic deformation (Hanmer et al., 1999a). Where the Big lake shear zone changes its trend from northwest to west (near 'A' in Fig. 2), 5–10 m wide MacQuoid dykes cut across granitic ultramylonite, and retrogressed granulite mylonite. These dykes have ophitic plagioclase, and well preserved chilled margins. Delicate apophyses of the larger dykes show similar relationships. This observation constrains the age of the mylonitization in the western segment to before 2.19 Ga. The orientation of the dykes changes to the southeast, following the swing of the shear zone trend (southeast of Squiggly lake; Fig. 2), reflecting subsequent large-wavelength east-northeast-trending F_4 regional folds which affect these rocks (Hanmer et al., 1999b; Ryan et al., 1999).

DEFORMATIONAL STRUCTURES

Orientation data

Eight fabric domains (Fig. 2) are characterized by the orientation and relative strain state of the dominant, steeply dipping local foliation. Fabric orientations from each domain are represented as pole figures on Figure 4.

Domains 1, 2, and 3 record a consistent variation in foliation attitude from northwest trending and steeply southwest dipping, to west trending and steeply to moderately north dipping. The extension lineation in all three domains is remarkably consistently plunging shallowly to moderately to the northwest. Regional foliation in domain 4 trends west-northwest to north-northwest, varying steeply about the vertical. Shear foliation in domain 5 is the most consistent in the entire area, trending northwest and dipping steeply southwest. The fabrics here also exhibit the highest degree of strain (Fig. 3f). Though somewhat variable, extension lineation in domain 5 generally has a shallow plunge to the northwest. Fabric elements in domain 6 are distinct in orientation compared to fabric within the Big lake shear zone. The shear fabric in domain 7 trends broadly west to west-northwest, varying about the vertical, and the associated regional extension lineation is shallowly west-northwest plunging. All of the above described fabric elements differ from those in domain 8, where shear foliation is steeply north dipping and the extension lineation plunges generally shallowly east or west.

The similarity in orientation of the shear foliation and regional foliation in domains 1 to 3 suggests that they were developed during the same deformation. Tight to isoclinal folds of foliation occur at the east and west margins of the shear zone, and plunge parallel to the extension lineation within the shear zone. The effects of these folds are evident from the pole figures (Fig. 4). Poles to the foliations in domains 1 to 3 lie along pi-girdles, and the poles to the girdles coincides with the extension lineation. This suggests that the folds may be developing late in the shear history. The discordance between fabric orientations in the Big lake shear zone and domain 6 may indicate that the foliation within the Cross Bay complex predated the Big lake shear zone.

Shear-sense criteria

Unlike the rather simple record of dextral shear sense in the central and eastern segments of the Big lake shear zone, shear-sense indicators in the western segment are less consistent, where the high-grade mylonite units preserve both sinistral and dextral shear-sense indicators. The high-grade mylonite units have been locally reactivated at lower grade, and the shear zone is affected by east-northeast-trending regional-scale folds and a northwest-trending fault (Fig. 1). The most reliable shear-sense indicators in the high-grade rocks are winged porphyroclasts derived from a variety of stiff objects (garnet porphyroblasts; feldspar phenocrysts, Fig. 3g; boudined stiff layers). Shear-sense indicators are not common, and are dominated by sinistral asymmetry rather than dextral. Local quartz mylonite units, derived from quartz veins, record dextral S/C fabrics with a grain size on the order 10–50 μm. It is unlikely that such fabrics in quartz survived granulite-facies metamorphism, and they are thus interpreted here as fabrics formed during lower grade reactivation within the shear zone. Quigley (1999) also noted dextral fabrics in quartz mylonite layers and attributed them to a greenschist-facies reactivation of the shear zone. He also noted an example of weakly developed dextral S/C fabrics in a lamprophyre dyke that crosscuts the mylonite. These lamprophyre dykes are similar in composition to lamprophyric phases within the Squiggly lake pluton interpreted to have commingled with the ca. 1.83 Ga (W. Davis, unpub. data, 1999) monzogranitic magma (Sandeman et al., 2000). More than 20 examples of undeformed lamprophyre were observed crosscutting the mylonite units. They generally preserve pristine phlogopite, whereas some exhibit chlorite alteration. Thus, these rocks

did not experience thermal conditions greater than greenschist facies after ca. 1.83 Ga. Moreover, because the contacts of the Squiggly lake granite are not offset, the lower grade fabrics within the shear zone are probably not regionally significant.

It is unclear how the sinistral kinematic indicators in the granulite- to amphibolite-facies mylonite units relate to the dextral shear history recorded in the eastern and central segments of the shear zone. The western segment of the shear zone is internally isoclinally folded, and the entire western end essentially outlines a large wavelength regional-scale fold (Fig. 2) that plunges shallowly to moderately to the northwest (*see* domains 1 to 3; Fig. 4). It is possible that some of the sinistral kinematic indicators were observed on opposite limbs of these internal isoclinal folds that postdated dextral shear. Ongoing microstructural investigation should better characterize shear sense recorded throughout the shear zone.

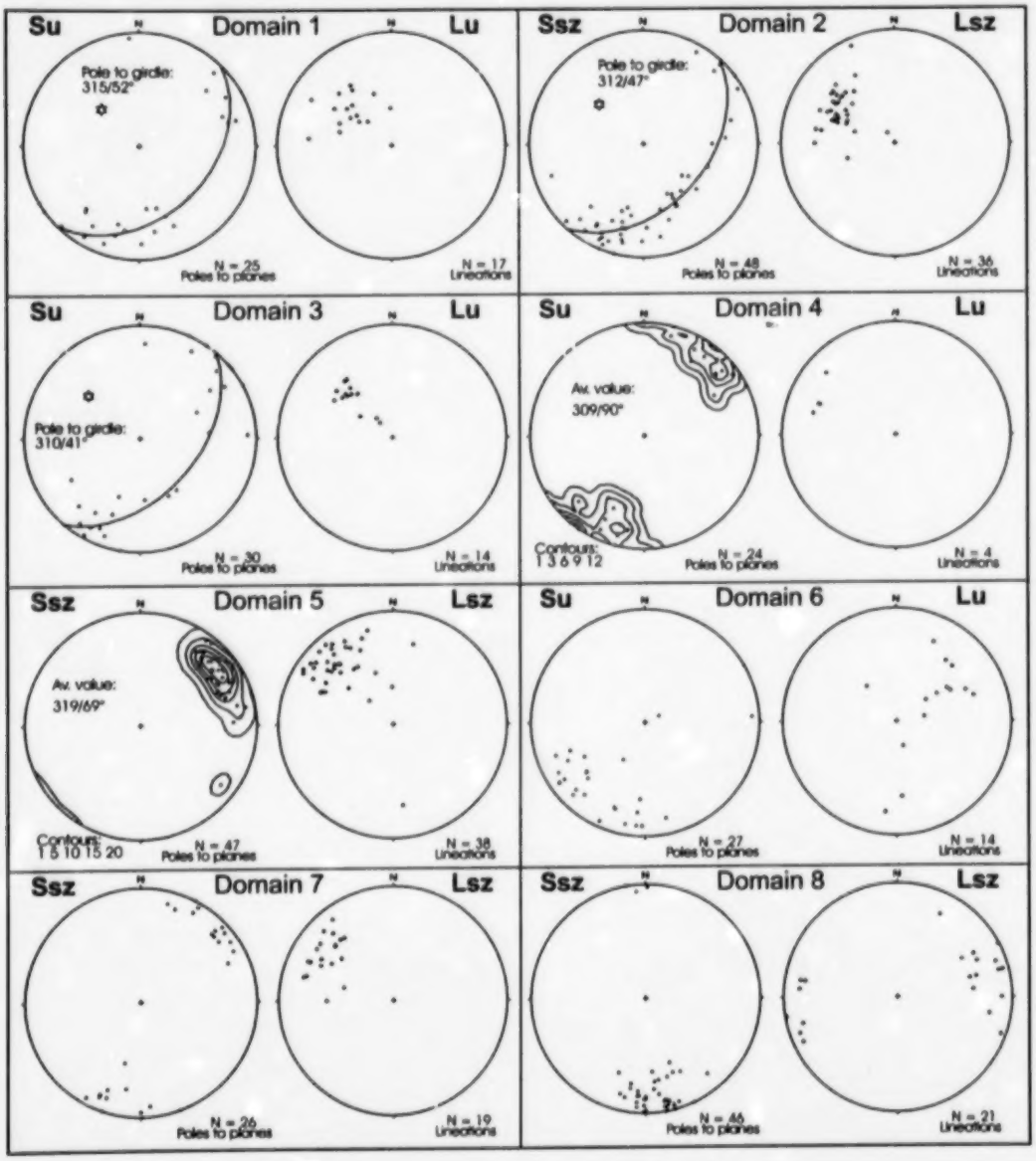


Figure 4. *Lower hemisphere equal area projections of planar and linear fabric elements from the eight structural domains outlined in the Squiggly lake area* (see *Fig. 2*).

Western termination of the Big Lake shear zone resolved?

The western end of the Big lake shear zone ends abruptly at a northwest-trending till-covered lineament, which is interpreted as a fault. Mylonite units of the Big lake shear zone were not observed on the west side of the lineament, and thus, field observations cannot demonstrate the sense of movement along this fault. Interpretation of regional aeromagnetic data (Fig. 5), and regional correlations of faults, elucidates why the shear zone ends so abruptly, and may provide clues to the origin of the Baker Lake Basin (*see* Rainbird et al., 1999).

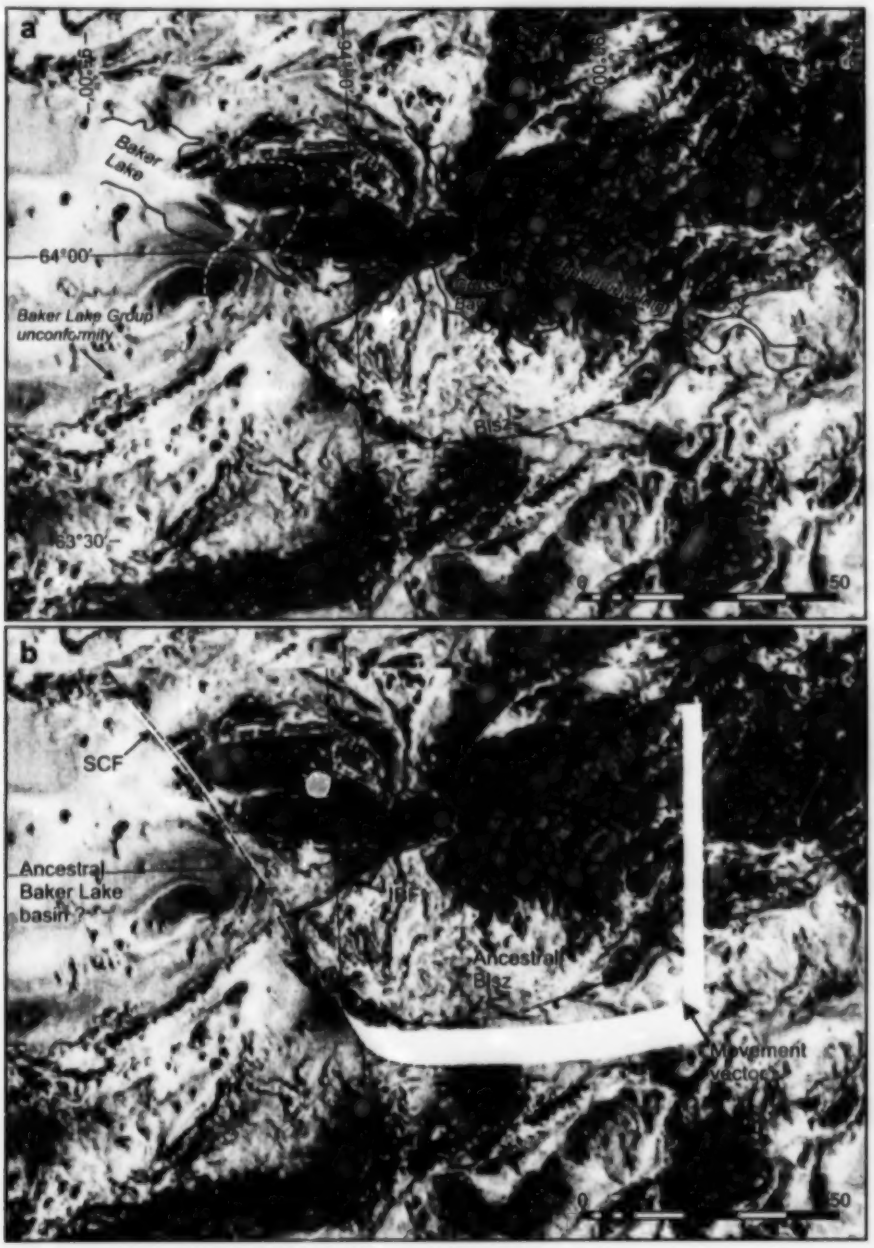


Figure 5. *Shaded total field aeromagnetic map of the study area, encompassing Gibson Lake–MacQuoid Lake–Baker Lake–Chesterfield Inlet area (illumination from the east).* ***a)*** *Original aeromagnetic map.* ***b)*** *Restored aeromagnetic map, resulting from displacing the northwest block by 10 km to the northwest. Blsz=Big lake shear zone, IBF= Ippijjuaq Bay fault, SCF=South Channel fault.*

Figure 5a illustrates that the western segment of the Big lake shear zone coincides with a northwest-trending aeromagnetic anomaly. In detail, it is not the Big lake shear zone that produces the anomaly, rather it appears to be magnetite-bearing amphibolite in the wall rocks of the shear zone. The northwest termination of the northwest-trending magnetic anomaly coincides with a northeast-trending magnetic anomaly (Fig. 5a), which steps 10 km to the southeast. We interpret the step in the northeast-trending anomaly to occur along a young northwest-trending fault which accordingly has about 10 km dextral separation. This proposed fault matches perfectly the orientation of a fault that cuts across Christopher Island and along South Channel in Baker Lake (Rainbird et al., 1999, and references therein). The fault offsets the basal unconformity of the Baker Lake Group by 10 km of dextral separation. It trends southeastward along South Channel into the MacQuoid–Gibson area (Fig. 1), and is thus informally termed the South Channel fault. The South Channel fault postdates the ca. 1.83 Ga Christopher Island Formation. Although the ca. 1.76 Ga Pitz Formation is not exposed near this fault strand, elsewhere in the Baker Lake Basin it is offset by faults of similar orientation and dextral separation. Thus, it is possible that the South Channel fault postdates 1.76 Ga.

In the aeromagnetic map, the trace of the South Channel fault coincides with the lineament where the Big lake shear zone terminates immediately west of Squiggly lake (Fig. 2, 5a), and we interpret this to be the same structure. Additional evidence for the fault lies in a chaotic disruption of the ultramylonite units along their southwesterly margin, south of Squiggly lake (Fig. 2). There, tight kink folds of the mylonite and the synkinematic mylonitized diabase units broadly have Z-asymmetry, and a chlorite-grade axial-plane crenulation cleavage. The trace of the fault is positioned west of this chaotic zone, at the edge of the narrow lake west of 'A' on Fig. 2.

To further illustrate this hypothesis, we restore the aeromagnetic data by moving the block on the northeast side of the South Channel fault by 10 km to the northwest along the trace of the fault (Fig. 5b). The result is a flat, wedge-shaped magnetic low, cored by a sharp magnetic high, that is distinct from the more complex high-amplitude magnetic pattern throughout the rest of the region. The northeast-trending anomaly (more appropriately, truncation of anomalies) on the northeast side of the South Channel fault, coincides with a 200–300 m wide fault and mylonite zone that is well exposed near Ippijjuaq Bay of Chesterfield Inlet; here informally termed the Ippijjuaq Bay fault. The Ippijjuaq Bay fault separates amphibolite-facies supracrustal rocks to the northwest from tonalitic and monzogranitic gneiss of the Cross Bay complex to the southeast. The main rocks types in the zone are red quartzofeldpathic mylonite, chlorite phyllonite and/or schist, and quartz-feldspar porphyroclastic cataclasite (apparently zones of brittle failure). The shear foliation in the Ippijjuaq Bay fault trends southwest, with moderate to steep dip, and has an extension lineation that plunges shallowly to moderately north-northeast. All shear-sense indicators, including asymmetric porphyroclasts, shear drag folds (Fig. 3h), and shear bands, are indicative of dextral shear, where the northwest side has down-dropped obliquely to the northeast.

Tella et al. (1999) mapped a steeply dipping, east-trending fault on Bowell Island (and west of the island) that separates metavolcanic and metasedimentary rocks on the south from high-strain tonalitic gneiss on the north (Fig. 1). The sense of displacement along this fault was not determined, but we suggest that the east-west fault and the Ippijjuaq Bay fault (Fig. 1, 5) may form a conjugate set, where the geometry is that of a down-dropped, wedge-shaped graben. Moreover, we interpret this graben and its bounding faults to represent an ancestral Baker Lake Basin. The gneiss units to the north, including the Kramanituar complex (Sanborn-Barrie, 1999), and the Cross Bay complex would have formed basement highs relative to the graben. The magnetic high in the central parts of the graben (which also coincides with a small gravity high) may represent a small, second-order horst, or merely the magnetic and gravity signature of a large gabbro body which intrudes the basement rocks at this locality (Fig. 1, 5a). Because clasts of the ca. 1.90 Ga Kramanituar complex (Sanborn-Barrie, 1999) occur in the South Channel conglomerate, above the unconformity (Rainbird et al., 1999), the basin-bounding faults are likely younger than 1.90 Ga.

DISCUSSION

Preservation of fine grain size in mylonite within the western segment (generally 10–50 μm), especially in the anorthositic and granulite-facies mylonitic rocks, is enigmatic. It probably indicates that these mylonite units cooled relatively quickly after deformation. The plagioclase-rich layers appear to have been particularly resistant to coarsening. The presence of radiating amphibole in some layers, as well as chlorite pseudomorphs after garnet, indicate the mylonite did interact with fluids after deformation ceased. However, the local preservation of granulite assemblages throughout the western segments indicates fluid infiltration was localized.

The anorthosite, granulite, and garnet-amphibolite bodies in the western segment of the shear zone bear a striking resemblance to rocks in the Kramanituar and Uvauk complexes exposed to the north of Chesterfield Inlet (Sanborn-Barrie, 1999; Mills et al., 1999; and references therein). Unlike the Kramanituar complex, however, which has been dated at 1.90 Ga (Sanborn-Barrie, 1999), the granulite units in the Big lake shear zone appear to be older than 2.19 Ga. Because no other documented regional tectonic events are recorded between 2.19 Ga and the Neoarchean, we interpret these mylonite units as Neoarchean. They are younger than ca. 2.69 Ga, the age of an augen granite deformed in the shear zone (W. Davis, unpub. data, 1999).

The granulite-facies mylonite units (and their variably retrogressed equivalents) within the western segment of the Big lake shear zone are thermally anomalous with respect to the amphibolite-facies mylonite and wall rocks. The spatial

association of the granulite, anorthositic sheets, and late-synkinematic diabase represents anomalous magmas, restricted to the western segment of the shear zone. We propose that the anomalous magmas were synkinematically localized in the western segment of the Big lake shear zone, and the residual heat from the anorthositic and mafic magmas induced the localized high-grade event. Synkinematic emplacement of the magmas helps explain why rocks that should normally be more competent than their tonalitic wall rocks, deformed so pervasively. We believe that thermal input from magma in a shear zone might be a general principle, and not a phenomenon restricted to the Big lake shear zone.

The steep attitude of the Big lake shear zone, with a shallow extension lineation, does not necessitate that the shear zone was a strike-slip structure. It is possible that the shear zone was originally a shallow mylonite zone that has been subsequently steepened by two regional folding events (F_3 and F_4 *in* Hanmer et al. (1999a) and Ryan et al. (1999)), which is consistent with the dispersion in orientation data in the western segment (domains 1 to 3 *in* Fig. 4). The Big lake shear zone now lies on the southern limb of the map-scale east-northeast-trending fold which controls the geometry of the Cross Bay complex. If the Big lake shear zone was originally a shallow structure, it is not clear whether the deformation regime would have been contractional or extensional.

While the significance of the northwest-trending fault that truncates the western segment of the Big lake shear zone was not readily apparent in the field, evaluation of regional aeromagnetic data and regional correlation of faults shed light on this issue. The Big lake shear zone appears to have been truncated first by the northeast-trending Ippijjuaq Bay fault, and then by the northwest-trending South Channel fault. The geometry and geographic location of the Ippijjuaq Bay fault, and related faults, are consistent with it being an ancestral basin-bounding fault to the Baker Lake Basin, and was likely active after 1.90 Ga. The Ippijjuaq Bay fault, the Christopher Island Formation, and the west end of the Big lake shear zone, have been offset by the South Channel fault, probably after 1.76 Ga.

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